DEVELOPMENT AND TESTING OF A WATER MANAGEMENT MODEL (WATRCOM): DEVELOPMENT

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ABSTRACT

Drainage is required in many agricultural watersheds in the southeastern United States for flood prevention and to sustain agricultural production. These drainage improvements often increase the severity of summer droughts by lowering water tables. A computer simulation model, WATRCOM, has been developed to assist in evaluating drainage improvements and the feasibility of using channel water level control. A finite element solution of the Boussinesq equation coupled with water balances in the unsaturated soil and on the surface is used to simulate water movement in three dimensions. Varying soil types and boundary conditions in land areas with irregular drainage channel networks can be considered. Model results are compared to published solutions for drainage to parallel drains. Solutions are also presented for flow in regions near intersecting drains and compared to solutions in regions with parallel drains. **KEYWORDS.** Drainage, Watershed, Channels, Hydrologic modeling, Water table, WATRCOM.

INTRODUCTION

Improvements of channel systems for drainage in relatively flat watersheds in the southeastern United States are important for flood control and management of seasonal high water tables. The improvements consist of deepening the channels and increasing drainage capacity. This may lead to overdrainage during the summer months. An alternative is the use of wide, shallow channels which will accommodate the surface drainage needs, but will provide very little subsurface drainage while requiring more land area.

The use of water control structures in deep channels allows management to insure drainage during times of excess water and storage of some of the excess water to reduce overdrainage during the summer months. Previous evaluations of the benefits of these structures have been limited to periods of excess water and to steady-state storage of water in the channel systems.

Approaches for describing saturated and unsaturated water movement on a watershed scale range from detailed numerical solutions (Freeze, 1971; Neuman, 1973) to models for the evaluation of land resource regions such as de Laat et al. (1981) and Querner (1984). Springer (1985) presented a review of several saturated-unsaturated flow models.

Freeze (1971) presented numerical methods to solve the three-dimensional Richard's equation for water movement in the saturated and unsaturated soils of a watershed. Applications of this approach were described by Freeze (1972a, 1972b). However, the input parameters and the computational requirements cause this approach to be expensive and difficult for the analysis of watershed scale systems.

Neuman's finite element model has been used for the analysis of two dimensional seepage. This model is an iterative Galerkin-type finite element solution of the two-dimensional saturated-unsaturated flow equations. The solution procedure can handle nonuniform flow regions with complex boundaries. Neuman (1973) analyzed transient saturated-unsaturated flow problems such as seepage through an earth dam and seepage through a layered hill slope cut by a ditch. The author limits his analyses to periods on the order of less than one month. The simulations are also limited to unsaturated-saturated responses to either fixed boundary conditions or constant fluxes on the boundaries. Field testing of the model was not presented.

The approach in Neuman's two-dimensional model has been extended to three dimensions by Huyakorn et al. (1986). They present six examples to verify and demonstrate the utility of their model in situations involving seepage faces and anisotropic media. Examples were presented for unsaturated-saturated responses to fixed boundary conditions or constant fluxes on the boundaries. For a steady flow problem (simulation of a 44-day period), the computer processor time was on the order of 177 CPU minutes on a DEC* VAX model 11/750. Field evaluations of the model were not presented.

The water management model, GELGAM, described by de Laat et al. (1981) has been applied to areas in the Netherlands for regional water resource planning. GELGAM is a distributed deterministic hydrologic model for the simulation of groundwater flow and evapotranspiration in large non-homogeneous areas. The

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model consists of components for saturated flow, unsaturated flow, and surface water flow. These components are coupled in time and the model provides estimates of hydrologic components such as groundwater elevations, runoff, and evapotranspiration on a regional basis.

FEMSAT is a finite element model for considering a saturated regional groundwater system (Querner, 1984). The model has been used to determine the relationship between phreatic surface and regional groundwater flow for input into other models. An example of this was presented by van Bakel (1986) using the simulation model SWADRE, a version of SWATRE (Belmans et al., 1983) and FEMSAT to evaluate subirrigation with open ditch drainage systems.

The implementations of the theory fall short in delivering a comprehensive simulation model that can be used to evaluate watershed-scale water management scenarios. The computer simulation model, WATRCOM, was developed for analyzing saturated and unsaturated water movement and storage in irregularly shaped drainage districts. Effects of multiple intersecting drainage channels, with and without channel water level control, can be analyzed for simulation periods of 1 year. This article describes the model and compares simulations for transient water movement to parallel drains to published solutions. The model was field tested using data from a watershed scale research project and these results are presented in a subsequent article.

Model Description

The WATRCOM model is based on a water balance in a region or element. The saturated portion of the model is simulated with a formulation of the two-dimensional Boussinesq Equation. The unsaturated zone in each element is a one-dimensional vertical water balance. The water balances are conducted on each time step and linked to each component by their respective boundary conditions. A surface water balance is also conducted at each time step. With these linkages and solution procedures, WATRCOM can simulate three-dimensional water movement in watersheds with shallow water tables. The WATRCOM water balance may be expressed for any time period, DT, as

$$\mathrm{DELSAT} + \mathrm{DELUNS} = \mathrm{RAIN} - \mathrm{AET} - \mathrm{OUTFLOW} - \mathrm{RO} - \mathrm{RSTOR} - \mathrm{PSTOR} \quad (1)$$

where

DELSAT = the change in the volume of water stored in the saturated zone per unit surface area (m),
DELUNS = the change in the volume of water stored in the unsaturated zone per unit surface area (m),
RAIN = the amount of rainfall per unit surface area (m),
AET = the actual evapotranspiration per unit surface,
OUTFLOW = the subsurface lateral flow across the boundaries per unit surface area (m),
RO = the amount of surface runoff from the region per unit surface area (m),
RSTOR = the change in potential runoff in retention storage per unit surface area (m),
PSTOR = the change in detention storage per unit surface area (m).

A flow diagram is presented in figure 1.

At the start of each day, rainfall intensity (breakpoint rainfall) and daily potential evapotranspiration (PET) are read from input files. The method for determining PET depends on the available weather data. The model selects the time steps based on the weather data. On days without rainfall, a time step of 4 h is used. A time step of 1 h is used for the periods with rainfall and 4 h for the remainder of the day on days with rainfall. The daily PET is distributed over time steps without rainfall and is assumed to be zero during periods with rainfall. At each time step, a

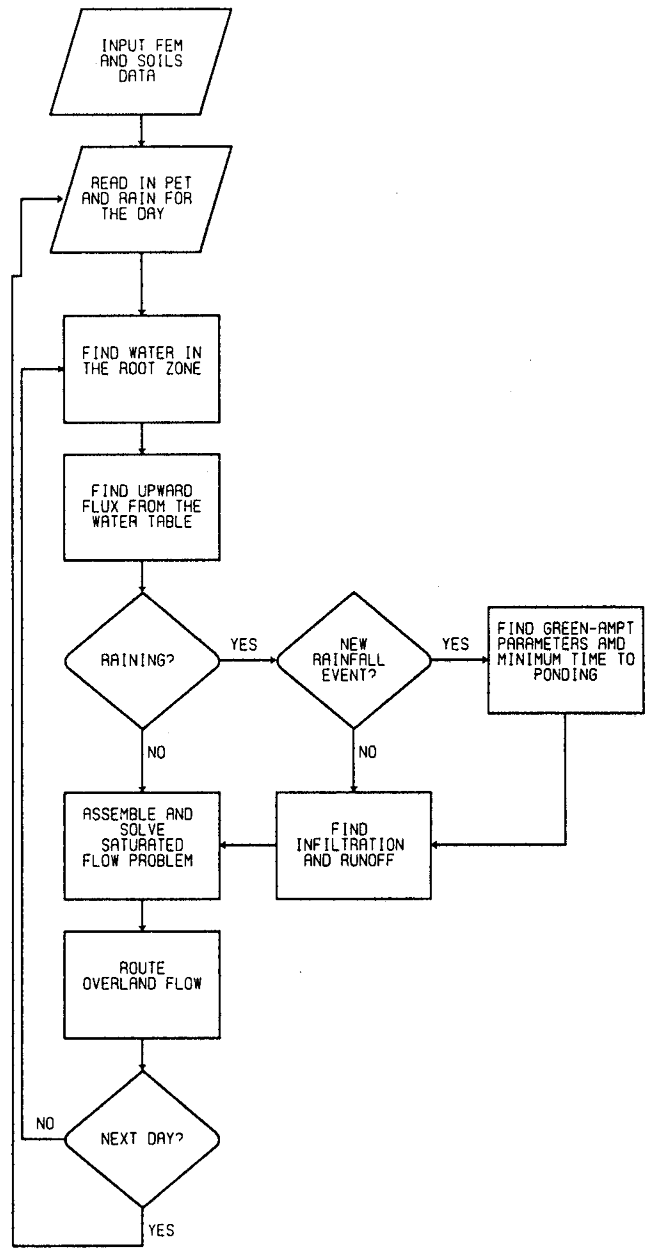

Figure 1–Flow chart of the simulation model, WATRCOM.

water balance is performed at each node by calculating the component terms in equation 1. The model consists of submodels for water movement in the saturated and unsaturated soil and overland flow on the soil surface. Descriptions of the calculation of the terms follow.

The Saturated Zone

The change in the water table elevation and the water stored in the saturated zone, DELSAT (eq. 1), is determined by solving the Boussinesq equation (van Schilfgaarde, 1974) for saturated flow. The equation may be written as:

$$f(h)\, h_t = \left(K(h)\, h\, h_x\right)_x + \left(K(h)\, h\, h_y\right)_y + R \qquad (2)$$

where

- h = water table height above the impermeable layer (m),
- $f(h)$ = drainable porosity, a function of h,
- $K(h)$ = lateral saturated hydraulic conductivity, a function of h (m/d),
- R = vertical recharge rate at the water table (m/d), positive for infiltration and negative for evapotranspiration,
- x, y = horizontal position coordinates of the region (m),
- t = time (days).

Equation 2 is based on the Dupuit-Forchheimer assumptions and neglects lateral water movement in the unsaturated zone. The drainable porosity, f, and the lateral saturated hydraulic conductivity, K, are functions of space and water table height. At each time step, the water balance in the unsaturated layers at each solution position is used to determine the vertical recharge rate, R, at the water table. Boundary conditions consist of zero horizontal flux on some boundaries and specified water levels versus time in the open channels throughout the area. If the channel becomes dry, the boundary condition may shift to one of zero flux. Figure 2 shows a two-dimensional schematic representation of the saturated portion of the model.

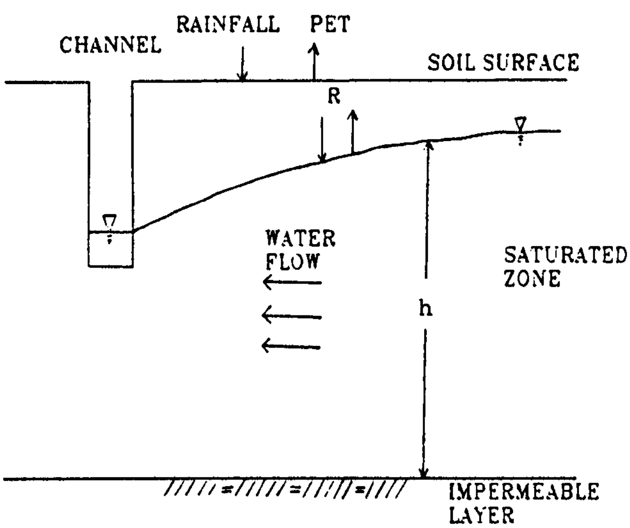

Figure 2–A schematic representation of the saturated portion of the model in two dimensions.

The solution procedure for equation 2 is the Galerkin finite element procedure with linear interpolation functions (White 1985). The region is divided into triangular elements (Norrie and de Vries, 1978). A system of nonlinear equations representing equation 2 is derived and the computations are simplified with two transformations. The first is the amount of drainable soil water in the profile, w:

$$w = \int_0^h f(h)\, dh \qquad (3)$$

The second transformation is the transmissivity function, T:

$$T = k(h)\, h \qquad (4)$$

where K(h) is dependent on water table elevation and may vary with depth and location in the region. Using equations 3 and 4, equation 2 becomes:

$$W_t = \left(T h_x\right)_x + \left(T h_y\right)_y + R \qquad (5)$$

Finite difference methods are used to express the time derivative, so at each time step equation 5 is a second-order partial differential equation in the space coordinates. This yields:

$$\left(w^{m+1} - w^m\right)/DT = \alpha\left[\left(T^{m+1} h_x^{m+1}\right)_x + \left(T^{m+1} h_y^{m+1}\right)_y + R^{m+1}\right] + (1-\alpha)\left[\left(T^m h_x^m\right)_x + \left(T^m h_y^m\right)_y + R^m\right] \qquad (6)$$

where

- m = the index associated with the time step,
- DT = time step (days),
- α = weighting factor between time step m and m + 1, selected between 0.67 and 1 for our problems ($\alpha = 1$ is fully implicit and $a = 0$ is explicit).

For any time step, the terms with index m are known. Equation 6 can be written with the unknown terms on the left and the known terms on the right as:

$$w^{m+1}/DT - \alpha\left[\left(T^{m+1} h_x^{m+1}\right)_x + \left(T^{m+1} h_y^{m+1}\right)_y\right] = w^m/DT + (1-\alpha)\left[\left(T^m h_x^m\right)_x + \left(T^m h_y^m\right)_y\right] + \alpha R^{m+1} + (1-\alpha) R^m \qquad (7)$$

A system of equations is assembled using the linear trial functions over each element in the simulation region. In matrix form, this system is (Parsons, 1987):

$$C\, w^{m+1} + \alpha\, DT\, BT_a^{m+1} h^{m+1} = r^m \qquad (8)$$

where

C = a sparse matrix of coefficients derived from the terms involving w in equation 7,

B = a sparse matrix of coefficients obtained from the h terms of equation 7,

T_a = a vector of the current estimates of the transmissivity at the adjacent nodes,

h = the vector of unknown nodal values,

r = the vector containing the known terms.

Newton's method is used to solve equation 8 for h at each node for each time step (Norrie and de Vrieg, 1978; White, 1985). The solution is accomplished by finding h for each node such that the function F(h) = 0 where F(h) is given by:

$$F(h) = C w^{m+1} + \alpha \, DT \, BT_a^{m+1} h^{m=1} - r^m \quad (9)$$

For Newton's procedure, an iteration sequence $h^0, h^1, ..., h^p$ is found as follows:

$$h^{j+1} = h^j + dh \quad (10)$$

where $dh = J^{-1} (h^{j+1}) F(h^j)$, and J is the Jacobian matrix at the unknowns h_{j+1}. A set of h's at each node is assumed to be those for which $F(h^{j+1})$ is sufficiently small. The K and f values at each iteration are based on the h found during that iteration.

THE UNSATURATED ZONE

The model performs a one-dimensional water balance in the vertical direction (Skaggs, 1978, 1980) at each node in the finite element grid. At each time step, approximate methods are used to predict the extraction of water from the profile for evapotranspiration (ET) and the addition of infiltrated rainfall. These balances provide the linkage to the Boussinesq equation (eq. 2).

At the start of the simulation, the soil profile at each node is assumed to be in hydrostatic equilibrium with the initial water table position. The soil profile is divided into 2 cm layers. The soil water content in each layer corresponds to a pressure head equal to the distance from the midpoint of the layer to the water table. When the water content is greater than the equilibrium amount, the excess is assumed to drain to the next layer. Water may be extracted from the root zone to satisfy ET requirements. However, the water content in the root zone layers cannot be lowered below the wilting point of the soil type for the node.

The balance for water extraction to satisfy ET at each node in the area can be written as:

$$AET_i = WSP_i + UPF_i + RZW_i \quad (11)$$

where

i = the node number in the area,

AET_i = actual evapotranspiration (cm),

WSP_i = amount evaporated from water ponded on the surface (cm),

UPF_i = water moving vertically from the water table to the root zone to meet ET (cm),

RZW_i = amount of water supplied from the root zone (cm).

Evapotranspiration is removed from the surface ponded water first, the water table second, and the root zone last. The amount of water moving from the water table vertically into the root zone, UPF_i, is found from the relationship between steady vertical upward flux and water table depth. The vertical upward flux versus water table relationship is derived from the unsaturated hydraulic conductivity function for each soil type (Skaggs, 1978, 1980). The vertical upward flux approximates the maximum amount of water that can move into the root zone in response to PET demands. If the PET is less than this amount, then AET and UPF are set equal to the PET amount and no water is extracted from the root zone.

If the PET is greater than UPF, then the difference between UPF and PET is extracted from the root zone water in a similar fashion as Skaggs (1978, 1980). The extraction takes place layer by layer from the soil surface downward. Water is extracted from each layer until the available amount in the layer reaches the lower limit. The procedure stops when all the water in the root zone has been depleted or the PET demand is met. AET will be less than PET creating a deficit when the root zone water is depleted.

The UPF term in equation 11 couples the water extraction routines with the saturated portion of the model at each node i. As the water table becomes deeper, less water will move into the root zone to meet the evaporative demand. The water available in the root zone will decrease, since the hydraulic head associated with the drained to equilibrium water content will decrease as the distance from the root zone to the water table increases. The term UPF is the amount of water moving from the water table to the root zone during the time step and the recharge term, R, in equation 2 is UPF_i/DT for node i.

INFILTRATION

In this model, a rainfall event is assumed to extend from the time rainfall starts until rainfall has stopped and all water on the surface has either infiltrated or runs off the area. The amount of infiltration at each node i is determined using the Green-Ampt equation with the assumption that the parameters are a function of the water table depth and soil type (Green and Ampt, 1911; Mein and Larson, 1973; Brakensiek, 1977; and Skaggs, 1980). The equation is given by:

$$INF_i = A_i / F_i + B_i \quad (12)$$

where

i = node number,

INF_i = infiltration rate (cm/d),

A_i = coefficient derived from the soil properties and the initial soil water content (cm /d),

B_i = coefficient derived from the soil properties (cm/d),

F_i = cumulative infiltrated water (cm).

The soil types and soil water contents at each node are used to determine the coefficients A and B by similar procedures described by Skaggs (1980).

Infiltration at each node is computed by first estimating the amount of water which can be infiltrated before ponding. The computed rainfall intensity, RFI_i, is compared to the Green-Ampt parameter for the node, B_i. If RFI_i is less than B_i, then the rainfall for this time step can be infiltrated. Otherwise, the amount of infiltrated water at ponding is given by:

$$INF_i = A_i / (RFI_i - B_i) \tag{13}$$

For the remainder of the first time step and each succeeding step in the rainfall event, the infiltration rate is computed using equation 12. The amount of time required to infiltrate a small amount, DR_i, is found. DR_i is added to the cumulative infiltration, F_i, and a new infiltration rate, INF_i, is computed using equation 12. At the end of the time step, a water balance is computed at the soil surface as:

$$RAIN = SUR_i + INF_i \tag{14}$$

where SUR_i is the amount of water (cm) added to the soil surface storage during the time step. The water added to the surface storage, SUR_i, is distributed as retention storage and detention storage which is eventually surface runoff, discussed below. The infiltrated water is added to the root zone layer by layer until the drained to equilibrium amount for each layer are reached. Once all layers are at drained to equilibrium, the remaining infiltration is assumed to move to the water table as vertical recharge during the time step. The amount moving to the water table provides the linkage between the infiltration water balance and the Boussinesq equation (eq. 2).

Overland Flow

The water on the surface, SUR_i (eq. 14), is distributed as retention and detention storage. Retention storage is the surface depression storage which will be infiltrated or evaporated. Detention storage is the water in excess of retention storage which may move to surrounding nodes. This water can leave the area via the boundary of the flow domain as runoff (RO in eq. 1). At the end of each time step, the water left in retention storage for each node is PSTOR (eq. 1). The water left in detention storage is RSTOR (eq. 1). At each time step, the water in detention storage is distributed over the area based on the slope of the water in detention storage between adjacent nodes. Runoff from the area is the water arriving at the boundary nodes during the time step.

Lateral Subsurface Discharge

Subsurface lateral outflow from the region is computed along each boundary element. Figure 3 shows a section of a finite element grid along a boundary. The estimate of the flow perpendicular to boundary side of each element is computed for each channel section in the area. The perpendicular from the interior node, k, to the boundary side ij is assumed to be the flow path for subsurface discharge. The equation to compute the discharge in m^3 along each boundary element is:

$$Q = -(K_a h_a (h_p - h_k) / L) \, b \, DT \tag{15}$$

where

Ka, ha = average saturated hydraulic conductivity (m/d) and the average water table elevation along the line perpendicular to the boundary side,

hp, hk = water table elevations on the boundary side and in the interior of the element (m),

L = distance from the boundary side to the interior node of the element,

b = length of the boundary side (m),

DT = time step (d).

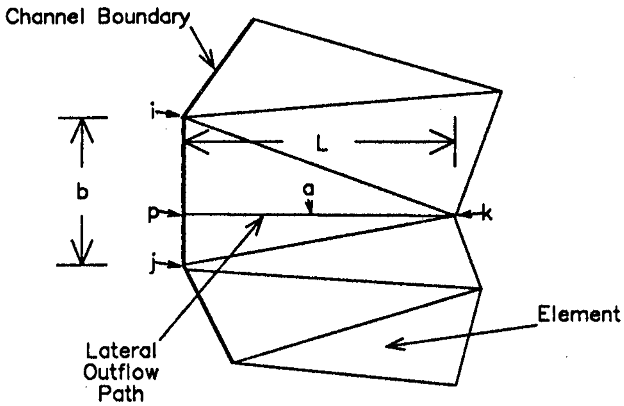

Figure 3–A diagram of the parameters used to compute the subsurface lateral flow at the boundary.

These discharges are summed along each boundary for lateral outflow from the region.

Water Storage and Availability

Water stored in the profile at any time is estimated as the amount of shallow groundwater available for drainage above the channel bottom. Each node in the area is assigned a boundary channel node by considering the transect perpendicular to the boundary section. The drainable storage is computed for each node at the end of each simulation day. The reference elevation used to compute the water storage at each node is the main drainage channel bottom elevation. The drainable water storage, S, at each node, i, is computed similarly to Badr (1983) as:

$$S = \int_e^h f(h)\, dh, \quad h > e \tag{16}$$

where

h = water table elevation at node i (m),

e = channel bottom elevation on the transect containing node i (m),

f(h) = drainable porosity function at node i.

Relative Yields

The model calculates wet and dry stress-day indices at each node in the region using procedures presented by Hardjoamidjojo and Skaggs (1982) and Evans et al. (1986). Inputs specify the planting and maturity dates. The total wet and dry stresses for the growing season are calculated and related to corn yield using the model developed by Hardjoamidjojo et al. (1982) for wet stresses and the model developed by Shaw (1978) for deficient soil water conditions.

Model Verification

Results predicted by the WATRCOM for transient water movement to parallel drains were compared to numerical solutions for drainage (Skaggs, 1973, 1976). This provides a validation of the saturated portion of the WATRCOM. Skaggs started with the formulation of the Boussinesq equation given by:

$$f h_t = K(h h_x)_x - R \qquad (17)$$

where
- h = water table elevation,
- x = spatial coordinate,
- t = time,
- f = drainable porosity,
- K = lateral saturated hydraulic conductivity,
- R = vertical recharge rate.

The boundary and initial conditions are written as:

$$h = h_d,\ x = 0 \text{ and } x = L,\ t > 0 \qquad (18a)$$

$$h = h_0,\ 0 \le x \le L,\ t = 0 \qquad (18b)$$

where
- h_d = elevation of the boundaries at the drains,
- h = initial water table elevation between the drains.

This corresponds to the one-dimensional version of the saturated portion of WATRCOM (eq. 2). The nondimensional form of equation 17 is more general than the dimensional form and was used for these tests. Equation 17 may be written in nondimensional form (Skaggs, 1973) as:

$$H_\tau = (H H_\beta)_\beta - \mu \qquad (19)$$

where
- H = h/hd,
- β = x/L,
- μ = RL^2/Kh_d^2,
- τ = $(Kh_d/fL^2)t$, and
- L = drain spacing.

The nondimensional forms of the boundary and initial conditions are:

$$H = 1, \beta = 0 \text{ and } \beta = 1, \tau > 0 \qquad (20a)$$

$$H = D,\ 0 \le \beta \le 1, \tau = 0 \qquad (20b)$$

where $D = h_d / h_0$.

The input parameters for WATRCOM were selected to obtain nondimensional solutions. The unsaturated portion of WATRCOM was turned off for these tests. WATRCOM used a nondimensional time step of 0.00208. The nondimensional time step for the finite difference numerical solution of equation 19 was initially 0.001. There are provisions to increase and decrease the time step as needed to ensure convergence of the numerical procedure as the simulation proceeded. The simulation period was 1 unit of nondimensional time. Simulations were conducted for boundary conditions corresponding to D = 0.2 and D = 0.8; the initial water table elevation was assumed to be at the surface in all cases.

Figure 4 shows the finite element grid used for the WATRCOM simulations. Table 1 presents the grid spacing sizes for each simulation. The methods are coded to indicate the solution procedure that was used. For example, ND-0.01 is the nondimensional finite difference method with a node spacing of 0.01, 3D-0.05 is the finite element model simulations for WATRCOM with a grid spacing of 0.05 in both the x and y directions, etc.

Predicted midpoint water table heights, H, of WATRCOM were in close agreement with the finite difference predictions, method ND-0.01, for both grid spacings. The largest difference in predicted midpoint H values was less than 0.01. Table 2 shows the means along with the root mean square error and the Pearson correlation coefficient between the finite difference solutions and the finite element solutions. The root mean square error (RMSE) and the Pearson correlation coefficient (CORR) are computed as:

$$\text{RMSE} = \left(\sum_{i=1}^{N} [\text{FDMH}_i - \text{FEMH}_i]^2 \right) / N \qquad (21)$$

and

$$\text{CORR} = \sum_{i=1}^{N} [(\text{FEMH}_i - \text{AFEMH})(\text{FDMH}_i - \text{AFDMH})] / (S_o S_s N) \qquad (22)$$

where
- N = number of nondimensional time steps,
- FDMH_i = nondimensional H simulated with the finite difference method at time step i,
- FEMH_i = nondimensional H simulated with the finite element method at time step i,
- AFDMH = average nondimensional H simulated with the finite difference method,
- AFEMH = average nondimensional H simulated with the finite element method,
- s_o = standard deviation of the nondimensional H simulated with the finite difference method,

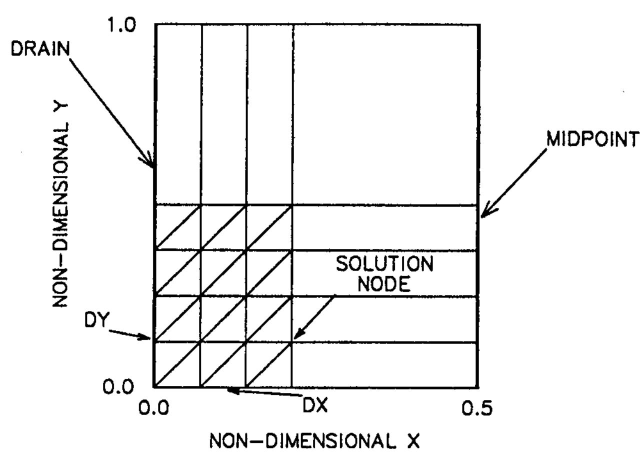

Figure 4–The finite element grid for the WATRCOM simulations of nondimensional water table heights.

TABLE 1. Nondimensional grid information for comparisons of the three-dimensional model to the finite difference approximations

Method	Dimension	Model	DX	DY
ND-0.01	Two	Finite difference	0.01	na*
3D-0.025	Three	Finite element	0.025	0.025
3D-0.05	Three	Finite element	0.05	0.05

* Not applicable.
(Source: Skaggs, 1973)

s_s = standard deviation of the nondimensional H simulated with the finite element method.

There was little difference in the predicted flowrates between any of the solutions for nondimensional time greater than 0.1. The largest differences in the predicted discharge rates occurred for $\tau < 0.1$. The smallest grid spacing size, method 3D-0.025, yielded the largest discharge rate for small times. The differences were on the order of 20%. Table 2 presents the mean discharge rates, the root mean square errors, and the correlation coefficients for the nondimensional Q comparisons. The coarser grid spacing, method 3D-0.05, predicted smaller mean discharges than the finer spacings during the early portion of the simulations for both boundary conditions. However, during the earlier portions, the simulated discharges from the WATRCOM solution were still larger than the finite difference solutions. For the simulation, the differences between the finite difference and WATRCOM predictions of discharge were less than 4%.

Example Application: Analysis of Water Movement Near Intersecting Drainage Ditches

Analysis of the performance of controlled drainage systems is difficult with current models. A typical section of a watershed in eastern North Carolina was selected to evaluate the use of WATRCOM characterize the effect of controlled drainage in fields close to uncontrolled collectors or main canals. The field ditches are 400 m long and spaced 100 m apart. Water control structures were located at the intersection of the field ditches and the collector canal. A schematic is shown in figure 5. The weir level or outlet elevation in the control structures was set at a depth of 0.6 m below the soil surface on day 1 of the simulation. Water was assumed to be pumped into the ditches to maintain the water level at the weir elevation. The collector was uncontrolled and the water level was assumed to be 2.0 m below the soil surface. The initial water table was assumed to be 2.0 m below the surface. The soil is a Rains sandy loam underlain by an impermeable layer at a depth of 4.0 m. Lateral hydraulic saturated conductivity is 5.0 m/d and drainable porosity is assumed to be constant at 0.1. The unsaturated data required for the model simulations are given by Skaggs (1980). WATRCOM was used to simulate this scenario using 1984 weather data from North Carolina. The weather data are summarized by months in Table 3.

Midpoint water table depths at distances of 25 m and 250 m from the collector ditch are plotted along with the daily rainfall distribution in figure 6. Drainage to the collector ditch influenced the water table response to the controlled field ditches. The water table responded to rainfall and ET throughout the year at both locations. However, drainage to the collector ditch caused the water tables near the collector ditch to remain approximately 0.6 m below those simulated at the 250 m transect. A two-dimensional analysis can be used reliably to determine water table response to controlled drainage or subirrigation at transects far removed from the collector ditch. However, seepage to the collector ditch is significant; it influences water tables close to the collector and the amount of water that must be pumped to maintain the ditch water level. Prediction of the behavior of the system near the intersection of the ditches requires a three-dimensional analysis which can be accomplished by WATRCOM.

Summary and Conclusions

A water management model, WATRCOM, for watershed scale drainage systems was developed. The model can be used to analyze the effect of channel water level control on soil water conditions and water conservation in drained agricultural watersheds. The model is based on water balances in subregions of the watershed. Components of the model consist of a finite element solution of the Boussinesq equation to characterize water movement in the saturated zone, a one-dimensional analysis in the unsaturated zone at each node in the finite

TABLE 2. Nondimensional H and Q comparisons of the finite element grid spacings for the three-dimensional model to the finite difference approximations*

	D = 0.2			D = 0.8		
Method	Mean	RMSE†	Corr‡	Mean	RMSE†	Corr‡
			Nondimensional H			
ND-0.01	0.431	-	-	0.828	-	-
3D-0.025	0.436	0.0023	0.998	0.829	0.001	0.998
3D-0.05	0.436	0.002	0.998	0.829	0.001	0.998
			Nondimensional Q			
ND-0.01	0.365	-	-	0.092	-	-
3D-0.025	0.367	0.009	0.998	0.096	0.030	0.994
3D-0.05	0.358	0.069	0.995	0.093	0.006	0.998

* See Table 1 for descriptions of methods. Source: Skaggs, 1973.
† Root mean square error between finite element simulated nondimensional H and finite difference solutions.
‡ Pearson correlation coefficient between finite element and finite difference simulated nondimensional H.

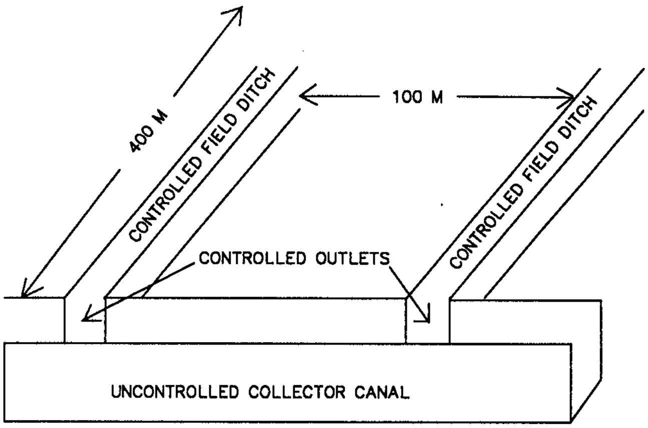

Figure 5–Schematic of controlled field ditches emptying in an uncontrolled collector canal.

TABLE 3. Monthly summary of the 1984 weather data

Month	Pet	Rain
	----------mm----------	
1	15	63
2	37	134
3	56	118
4	79	88
5	136	183
6	172	58
7	136	250
8	145	87
9	96	124
10	71	8
11	32	24
12	27	37
Total	1002	1174

runoff. All components are coupled at each time step.

The model simulation procedures were tested using published solutions for parallel drainage to open ditches. WATRCOM predictions of water table height were within 1% of the published finite difference solutions for all cases considered. Water table height predictions showed little sensitivity to grid spacing size. Predicted discharge rates were acceptable with some deviations from published solutions occurring during the early portion of the simulations. Differences in discharge rates over the simulation were less than 4%. Finer finite element grid spacings tended to increase predicted discharge rates. These discharge rates were larger than published solutions obtained by finite difference methods during the early part of the drainage event.

Water table response near the intersection of field and collector ditches was simulated for one year of climatological record to illustrate the need for a three-dimensional analysis in some watershed problems. The water table in areas of the field close to the collector ditch was influenced by both the field and collector ditches making a two-dimensional analysis unsatisfactory for this situation.

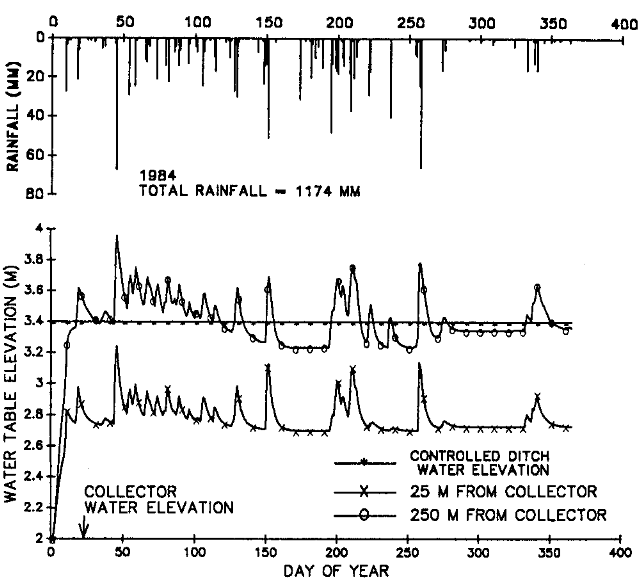

Figure 6–A comparison of midpoint water table response to controlled field drainage ditches intersecting an uncontrolled collector ditch.

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